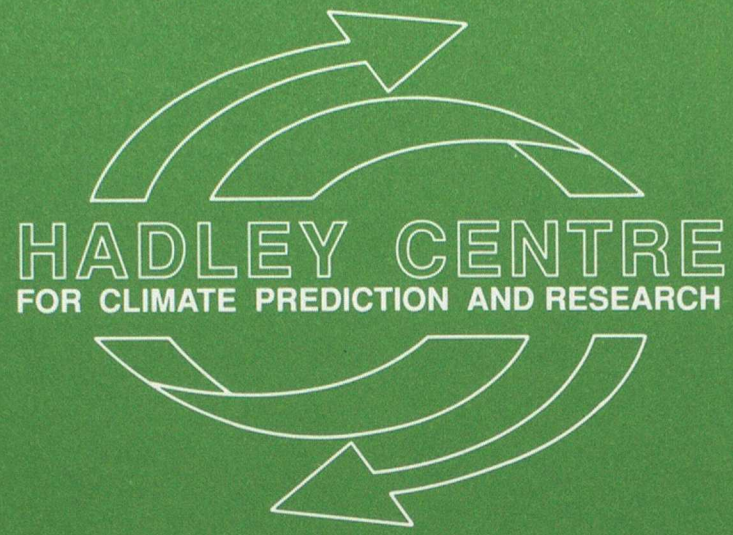


Modelling of Palaeoclimates:

Examples from the recent past

by

J. F. B. Mitchell

CRTN 37 April 1993

CLIMATE
RESEARCH
TECHNICAL
NOTE

Hadley Centre
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY

CLIMATE RESEARCH TECHNICAL NOTE NO. 37

MODELLING OF PALAEOCLIMATES:

EXAMPLES FROM THE RECENT PAST

by

J F B Mitchell

Paper presented at the Royal Society meeting on
"Palaeoclimates and their modelling with special reference to
the Mesozoic era", London, 24th and 25th February, 1993.
A shortened version will appear in the Proceedings of the meeting.

Hadley Centre for Climate Prediction and Research
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY

NOTE: This paper has not been published. Permission to quote from it
should be obtained from the Director of the Hadley Centre.

Modelling of palaeoclimates: Examples from the recent past

JOHN F.B. MITCHELL

Hadley Centre for Climate Prediction and Research, Meteorological Office, Bracknell, RG12 2SY, U.K.

KEYWORDS

Holocene, initiation of ice sheets, last glacial maximum, general circulation models, orbital variations, carbon dioxide.

SUMMARY

Three dimensional general circulation models have been used in equilibrium studies of past climates by fixing the slowly changing parts of the climate system (orbital parameters, ice sheets, atmospheric composition). Results are presented from studies of the mid-Holocene, the last glacial maximum and the initiation of the last ice age.

1. INTRODUCTION

Over the last two and a half decades, advances in computing power have made possible the development of detailed three-dimensional global models of climate (general circulation models, or GCMs) . These models are based on physical principles (see, for example, Schlesinger,1988; Trenberth, 1992), have been validated extensively against present climate, and have shown some skill in reproducing the main features of the general circulation of the atmosphere (Gates et al, 1990,1992). Averaged over areas of $10^6 km^2$, typical seasonal mean errors are a few degrees Celsius in surface temperature, and several tens of percent in precipitation (Gates et al, 1990). The validation of ocean GCMs is less complete, due partly to there being less observational data available, and the inherently smaller scale of synoptic disturbances in the oceans (10-100 km as opposed to 1000 km in the atmosphere).

The main application of such models has been in attempts to understand and predict the climatic effects of increases in greenhouse gases (for example, Mitchell,1989; Houghton et al,1990,1992). Although there is broad agreement between different models on the nature of changes at the largest scales, there is poor agreement on regional scales, and the response of the models is often dependent on the parametrization of processes not resolved on the models' grid. In fact, two models can produce similar simulations of present climate, but substantially different simulations of the response to increases in greenhouse gases (Senior and Mitchell, 1993), so validation against present climate is not sufficient to discriminate between reliable and unreliable models.

General circulation models have been applied extensively to past climates (see, for example, reviews by Hecht, 1985; Saltzman,1990). Much palaeoclimatic data is obtained from cores from which one can reconstruct the climatic history back through different geological times at a given site. At present, the length of GCM experiments is limited to decades by the availability of computer resources, so we cannot yet attempt to simulate the evolution of climate over geological timescales . Instead, it is generally assumed that on short time scales (hundreds of years) climate is in approximate equilibrium which is determined uniquely by the state of the slower changing components of the system, such as the major ice sheets, atmospheric composition and the earth's orbital parameters. These components, or "boundary conditions", are prescribed and the response of the faster changing components of climate, such as cloud, sea-ice and ground wetness are simulated by the model. This approach gives "snapshots" of specific times in the past. Current models still omit some of the fast processes - for example the effects of vegetation, atmospheric dust and aerosols are not yet included interactively, and the sea surface temperatures are usually prescribed or determined by a simple thermodynamic mixed-layer model of the upper ocean. It is assumed that the final equilibrium is independent of the initial conditions. This is generally true for atmospheric models, but not always true for coupled ocean atmosphere models (Manabe and Stouffer,1988).

Periods for which the cause of climate change is well known, and for which there is wide geographical distribution of well-dated and accurately calibrated palaeodata, can be used for model validation. Considerable progress has been made in producing synoptic data for specific periods from the last glacial maximum to present (see for example, Climate Long Range Investigation Mapping and Prediction project members, 1976,1981 - henceforth referred to as CLIMAP 1976,1981- and Co-operative Holocene Mapping Project (COHMAP) members,1988). Note however that there are

problems of interpretation when comparing model grid values, which are representative of areas of 10^5 km, with observational data from specific sites, even with instrumental data for our present climate (Reed, 1986). This problem is compounded because of the uncertainties in estimating changes in temperature and precipitation from palaeoclimatic data.

The ability of a model to simulate a past climate does not necessarily guarantee that it will faithfully predict the effect of increases in greenhouse gases as the mechanisms of change may be different (Mitchell,1990). However, failure to simulate a past climate successfully may help to identify shortcomings in model formulation.

If the changes in boundary conditions are well established, the model can be used to determine the accompanying changes in atmospheric circulation, frequency of storms and other parameters not easily derived from field data. Some indication of the reliability of the model can be obtained by comparing the simulated changes in temperature and precipitation with reconstructed data, in regions where it is available. If such a comparison is favourable, the simulated data can also be used to estimate changes in regions where palaeoclimatic data is not available. This approach has been used, for example, in helping to reconstruct the climate of the last glacial maximum, as discussed later (Section 4).

For periods for which the changes in boundary conditions are not well established, one can investigate the effect of changing different boundary conditions singly or in different combinations. By comparing the simulated and reconstructed climates, one may gain some insight into the causes of the change in climate, and the accompanying changes in circulation. Model simulations may help to highlight where competing theories of change diverge. This would allow one to optimise the choice of site to obtain data to discriminate between different hypotheses. Examples of the effect of changing different boundary conditions will be given for both the mid-Holocene and the last glacial maximum.

In principle, one can use the "snapshot" approach to determine whether or not the simulated climate is in long term equilibrium with the prescribed boundary conditions. For example, if there is a net annual increase in snowdepth in high latitudes over land, one might conclude that this would lead over thousands of years to the formation of an ice sheet in that region. This technique has been used in attempts to determine what initiated the last ice age (Section 5).

Given the inaccuracy of present models on regional scales, such an approach must be regarded as speculative. On the other hand, if one can determine the rate of change of the slowly changing boundary conditions from equilibrium experiments in which the boundary conditions are prescribed, it may be possible to parametrize those rates of change as a function of the boundary conditions themselves. One could then model the evolution of the slowly varying components of climate in a much simplified model. Saltzmann (1990) gives a more detailed discussion of the philosophy and problems of modelling past climates.

2 MODELS

In atmospheric GCMs, prognostic equations governing the conservation of momentum, heat, mass and water substance are solved numerically over the globe at a set of discrete levels in the vertical. These equations have the general form

$$\frac{\partial Y_i}{\partial t} = \underline{\nabla}.(\underline{V} Y_i) + P + S + D \qquad (1)$$

where Y_i represents the east-west and north-south components of velocity, temperature and specific humidity and $\underline{V}$ is the horizontal velocity. P represents the effect of pressure gradient (adiabatic vertical motion) in the velocity (temperature) equations and D represents smoothing terms necessary for numerical stability. S represents local sources and sinks including those due to processes not resolved on the numerical grid. The largest source of uncertainty is generally associated with the parametrization of the these sub-gridscale processes (for example, cloud, radiation, cumulus convection, surface evaporation and surface drag). For some processes (for example, radiative transfer) parametrizations are based on the underlying mathematical theory. The effect of the approximations necessary to keep the computation time down can be evaluated against the exact formulation. However, for many processes, no mathematical basis has been established, so parametrizations are developed from observational studies, laboratory experiments, or numerical experiments on a fine numerical mesh, as appropriate. Some tuning of disposable parameters may be necessary, using sensitivity experiments. In general, however, the degree of tuning is limited, and confined to global parameters.

Since the right hand side of (1) can be evaluated in terms of current variables, the equations can be stepped forward in time using finite difference or spectral techniques. Current atmospheric models use a horizontal grid which is typically 300km or 30

spectral waves for climate applications, and 10 to 20 vertical levels (see, for example, Table 1). The current Hadley Centre model requires 18 hours of elapsed time to simulate one year using two processors of a CRAY YMP. Atmospheric models can be validated by comparing the simulated statistics of present climate (means, variances, seasonal cycle) against observed climatologies, and by comparing their response to contemporary changes in boundary conditions (especially sea surface temperatures) against the observed response (see Gates et al, 1990,1992). Schlesinger (1988) and Trenberth (1992) give more detailed accounts of the construction and use of climate GCMs.

Ocean GCMs are constructed in a similar way to atmospheric models, except that salinity rather than specific humidity is used, and spectral methods of solution are less appropriate because of the irregular boundary of the continents. The parametrization of physical processes is generally simpler. In validating ocean models, considerable use has been made of tracer distributions, including those of tritium (Sarmiento, 1983),carbon isotopes (Toggweiler et al, 1988) and CFCs (Dixon et al, 1993).

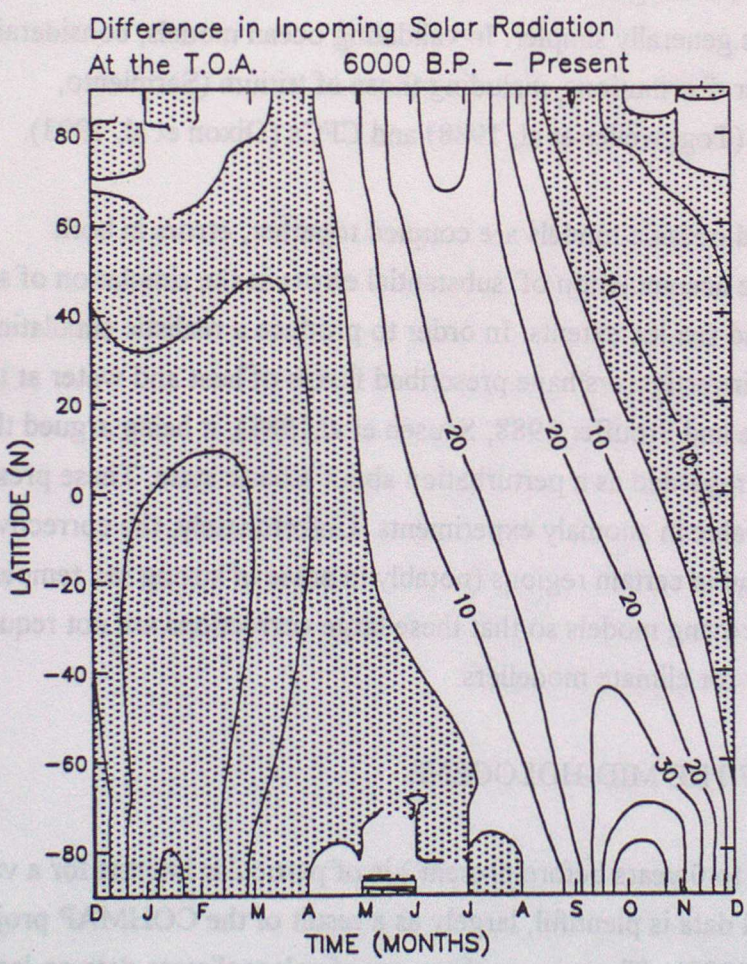
When atmospheric and oceanic models are coupled together, errors in both components lead to the accumulation of substantial errors in the simulation of sea surface temperature and sea-ice extents. In order to produce a realistic simulation of present climate, some investigators have prescribed fluxes of heat and water at the ocean surface (Manabe and Stouffer,1988, Sausen et al,1988), it being argued that climate change can be regarded as a perturbation about a basic state. These prescribed fluxes keep the same value in anomaly experiments. Unfortunately, the corrective fluxes required are large in certain regions (notably in areas of strong sea temperature gradients). Hence improving models so that these large corrections are not required remains a high priority for climate modellers.

3 SIMULATIONS OF THE MID-HOLOCENE

The Holocene (12000 to 0 years before present) is of particular interest for a variety of reasons. First, good data is plentiful, largely as a result of the COHMAP project (COHMAP members,1988). There is a wide range of palaeoclimate data on land at all latitudes, whereas such evidence from earlier periods in much of the extratropics has been destroyed by successive glaciations. Data from cores is more easily obtained than that for earlier periods which are generally more deeply buried. The errors in dating are also relatively small. Second, the known changes in boundary conditions were simple (see, for example, Kutzbach and Street-Perrott,1985). Changes in the earth's orbital

parameters (increased obliquity, and a change in perihelion from northern winter to northern summer) led to increases of up to 8% in northern hemisphere summer insolation, and decreases in winter, giving an enhancement of the seasonal cycle. The changes for 6000 years before present (6kBP) are shown in Figure 1. The concentration of greenhouse gases was similar to pre-industrial levels, and the estimates of sea surface temperature are similar to present (Ruddiman and MacIntyre, 1981). At 9kBP, there was still a substantial Laurentide ice-sheet, but by 6kBP this had virtually disappeared. Changes in atmospheric aerosol are thought to have been small, although there were some substantial changes in vegetation, particularly in the subtropics.

FIGURE 1



1. Changes in insolation at 6Kbp at the top of the atmosphere due to changes in the earths' orbital parameters. Contours every 10 Wm^{-2}, decreases stippled.

Table 1. GCM Simulations of the Holocene

Study	Period	Seasonal Cycle	Atmosphere Horizontal & Vertical Resolution	Ocean, or SSTs	Land Ice	Ice- free Land Albedo	Length of experiment
Mitchell ,1977	10k	Yes	3x3, L5	Present	Present	Present	60 days
Kutzbach and Otto Bliesner, 1982 (also Kutzbach,1981)	9k	Yes	R10,L5	Present	Present & 9K	Present	14 months
Kutzbach and Guetter, 1986 (also Prell & Kutzbach, 1987)	3,6k	January & July	R15, L9	Present	Present	Present	450 days (January)
	9k				9k		150 or 450
	12k*			12k	12k	12k	days (July)
Rind et al, 1986	11k	Yes	8x10, L9	Present &18k	Present &18k	Present	4 years
Kutzbach & Gallimore,1988	9k	Yes	R10,L5	50m ML	Present	Present	9 years+
Mitchell et al,1988 (see also Mitchell,1990)	9k	Yes	5x7.5, L11	50m ML	Present & 9K	Present	13 years+
Street-Perrott et al, 1990	9k	Yes	5x7.5, L11	50m ML	Present	9k Africa & Arabia	6 years+
Liao (1992)	6k	Yes	5x7.5, L11	50m ML	Present	Present	10 years+
Mitchell & Hewitt (this paper)	6, 9k	Yes	2.5x3.75,L19	Present	Present	Present	5 years
Hall (pers comm)	6k	Yes	T31, L19	Present	Present	Present	In progress

Resolution is either degrees (latitude by longitude) or number of spectral waves, R or T for rhomboidal or triangular truncation respectively. L indicates the number of levels. ML= mixed layer. k=1000 years before present.

* CO_2 levels were slightly reduced

+ does not include time to spin-up equilibrium

A large number of experiments have been carried for the mid-Holocene (Table 1). The complexity of the models and the length of simulations have increased with increasing computing power. Most simulations have been for 9kBP, when the changes in insolation were largest, or 6kBP, when the ice sheets were similar to present, and there is more data. Results for 6kBP are similar to those at 9kBP.Here, we summarize results for both periods (see Table 1), including figures from a simulation for 6kBP

using a preliminary version of a new Meteorological Office climate model (Cullen, 1993; Mitchell and Senior 1993). These findings are typical of parallel studies.

The increased insolation in the boreal summer (Figure 1) produces a general warming, especially in the northern hemisphere. The continental surface warms faster than that of the ocean (Figure 2a) because of the smaller thermal inertia (or because the ocean temperatures are prescribed- see Table 1). The enhanced warming of the atmospheric column over land leads to divergent flow in the upper troposphere, a lowering of surface pressure (Figure 2b), and stronger low level convergence from the oceans near the surface. This increases the precipitation associated with the northern hemisphere monsoons over India, the Sahel and Venezuela (Figure 2c). The ascent over the northern continents is accompanied by subsidence over the neighbouring oceans. In short, the enhancement of the seasonal cycle of radiation leads to more vigorous northern hemisphere monsoons.

The increased land sea contrast produces stronger monsoon flow over the northern Indian Ocean in summer. Luther et al (1990) and Bigg et al (1991) used surface winds from atmospheric model simulations to drive mesoscale ocean models of the Indian Ocean. Both found increased and more extensive upwelling in the Arabian sea in summer, consistent with palaeoclimatic indices (for example, Prell and Kutzbach,1987).

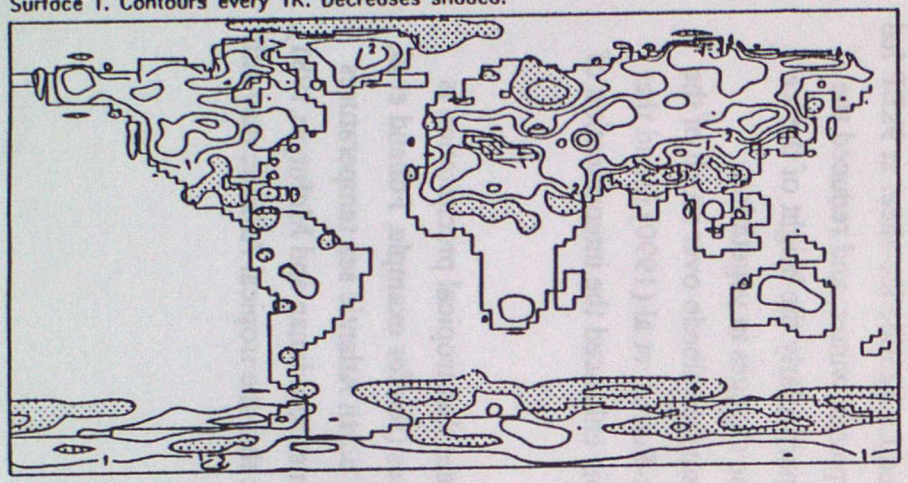
In northern winter, insolation is reduced, cooling the continents, increasing surface pressure in the tropics and reducing precipitation over tropical land (see, for example, Kutzbach and Guetter,1986; Mitchell et al, 1988). The ground wetness increases in the northern subtropics because of increased convergence and precipitation in summer (Figure 2d). In midlatitudes, soil moisture decreases (Gallimore and Kutzbach,1989; Mitchell,1990). Here, the increase in evaporation due to enhanced insolation in summer is greater than the increase in precipitation. Large scale precipitation in mid-latitudes is associated more with baroclinic eddies driven by the equator-to-pole gradient than with low level convergence resulting from land-sea temperature contrasts. The simulated changes in soil moisture are reasonably consistent with estimates of lake levels for the period (Street-Perrott and Roberts,1983).

In simulations with a simple ocean, the sea temperature changes lag the land changes by two months or so. The extra solar heat in summer is stored in the ocean and released in late autumn and early winter, delaying the onset of winter cooling over the northern continents.

FIGURE 2

(a)

Differences between 6kBP and present for JJA
Surface T. Contours every 1K. Decreases shaded.

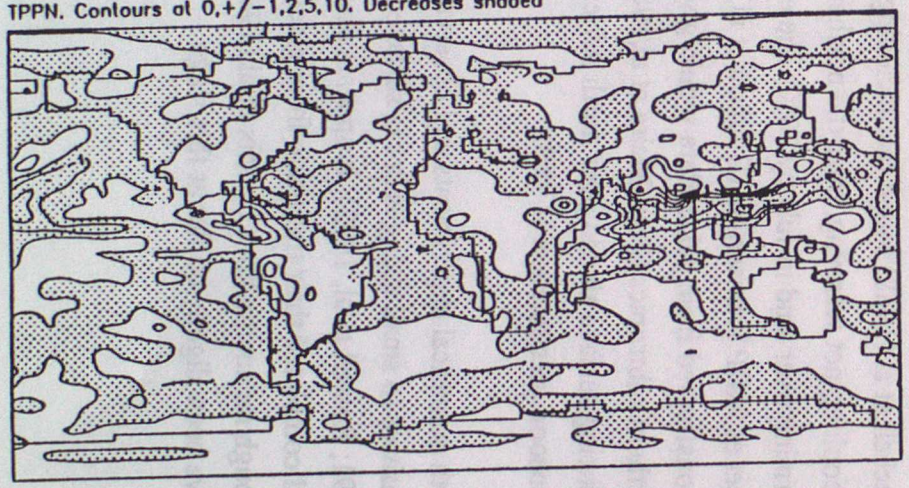


(b)

Differences between 6kBP and present for JJA.
PMSL. Contours at 0,+/−1,2,4,8mb. Decreases shaded

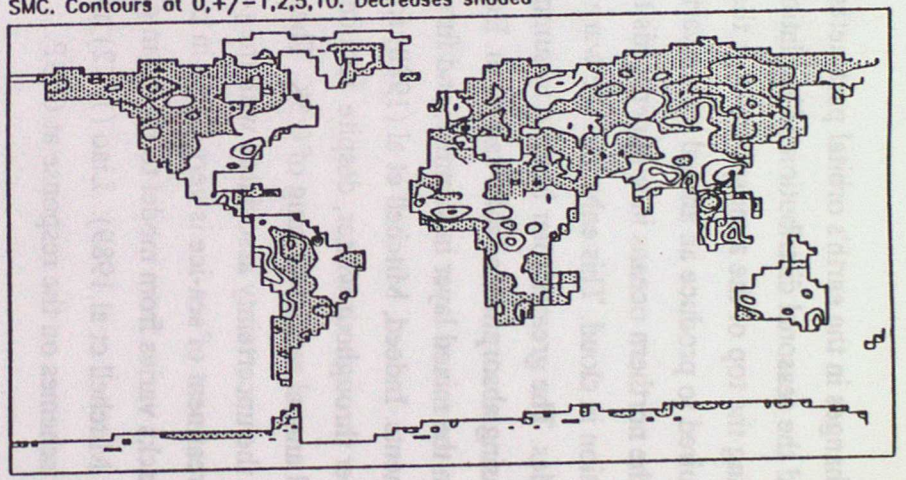
(c)

Differences between 6kBP and Present for JJA
TPPN. Contours at 0,+/−1,2,5,10. Decreases shaded



(d)

Differences between 6kBP and Present for JJA.
SMC. Contours at 0,+/−1,2,5,10. Decreases shaded



2. Changes (6kBP-present) during June,July August simulated using a preliminary version of the Meteorological Office unified forecast/climate model. (see Table 1).

The changes in the earth's orbital parameters over the last 125 thousand years have altered the seasonal distribution of insolation, but not the total amount of radiation reaching the top of the atmosphere over the year. Thus some non-linearity of response is required to produce an annual mean warming (or cooling). The increased subsidence over the northern oceans in summer leads to a drying of the marine atmosphere and a reduction in cloud. This enhances the warming of the oceans, particularly in mid-latitudes. The greater solar heating in summer reduces Arctic sea-ice extents, further increasing absorption of solar radiation. Both these feedbacks increase the storage of heat in the mixed layer in summer, and further delay the winter cooling over the continents. Indeed, Mitchell et al (1988) found that the northern extratropics were warmer throughout winter, despite reduced insolation in that season, and there was a global annual mean warming of 0.5K. These results must be regarded as tentative given the uncertainty associated with the parametrization of sub-grid scale processes. The treatment of sea-ice is very crude in current models, and even the sign of cloud feedbacks varies from model to model in global warming experiments (Cess et al, 1989; Mitchell et al,1989). Liao (1992) has investigated the influence of different cloud schemes on the response at 6kBP.

The simulated climate at 9kBP is sensitive to changes in other boundary conditions. Mitchell et al (1988) found that including an idealised Laurentide ice-sheet at 9kBP led to cooling, not warming, over the northern extratropics in winter, and reduced the warming over land in July by 20%. The extent and particularly the height of the ice sheet at 9kBP is uncertain (Boulton et al, 1985). The changes in vegetation reconstructed from palaeodata would give a lower surface albedo over much of the northern subtropics (Petit-Maire et al,1988). Street-Perrott et al (1990) found that reducing surface albedo over north Africa and Arabia enhanced the intensification of monsoon precipitation at 9kBP.

Both numerical experiments and observations indicate that tropical precipitation is sensitive to small changes in sea surface temperatures (see for example, Folland et al, 1991; Rowell et al, 1992). It appears that by 9kBP, north Atlantic sea temperatures had completed their transition from glacial conditions (Ruddiman and McIntyre,1981), though reconstructions by Mix et al (1986) suggest that the tropical Atlantic may still have been slightly cooler at this stage.

Finally, there have been attempts to use model simulations to explain the Younger Dryas, a temporary reversion to colder conditions at around 11-10kBP (Rind et al, 1986 - see also Manabe and Stouffer,1988)

4. SIMULATIONS OF THE LAST GLACIAL MAXIMUM

The climatic changes during the last glacial maximum (about 18kBP) were much larger than at 6 or 9kBP. However, the cause of the changes is not well established. The earth's orbital parameters, which appear to explain most of the differences in the middle Holocene, were fairly similar to present (Berger, 1978). The largest differences in the distribution of radiation occurred at the equinoxes, and changes in solsticial seasons in high latitudes were small. There were large ice sheets over North America and northwest Europe (Denton and Hughes,1981), sea temperatures were generally cooler, especially in the North Atlantic ocean, sea-ice was more extensive and snow-free land albedos were higher in much of the tropics and subtropics (CLIMAP, 1976, 1981). CO_2 levels were less than two thirds their present value (Barnola et al, 1987). It is likely that the atmospheric aerosol loading was higher than present, though this effect has yet to be represented in numerical simulations.

Most experiments to date have taken into account the ice-sheets over the northern continents, with the changes in sea surface temperatures and sea-ice cover either prescribed or calculated in the model (see Table 2). These changes lead to a simulated global mean cooling which, in general, is greater over land than over the neighbouring ocean. This produces higher surface pressure over much of the northern extratropical continents, and hence reduced precipitation (for example, Manabe & Broccoli,1985a; Kutzbach & Guetter, 1986; Rind,1987). The ice sheets displace the jet stream and the associated storm tracks southwards, particularly over North America, producing a marked reduction in precipitation immediately south of the ice sheets in most models. Westerly flow is also diverted around the north of the Laurentide ice sheet and then south between North America and Greenland. This cold southward outflow would have helped to maintain the eastern flank of the ice sheet, and cooled the northern Atlantic ocean. Even so, the simulated energy balance indicates that the Laurentide and Eurasian ice sheets would not be in equilibrium, but would melt, particularly at lower elevations (Manabe & Broccoli,1985a; Rind, 1987).

Broccoli and Manabe (1987) have estimated the separate effects of the changes in land-ice, CO_2 and vegetation (through surface albedo) using an atmospheric model coupled to an oceanic mixed-layer model. The effects of the changes in vegetation are

Table 2 GCM Simulations of the last glacial Maximum

Study	Seasonal cycle (period covered)	Atmospheric Horizontal & Vertical Resolution	Ocean or SSTs+	CO_2	Land Ice	Snow - Free Land Albedo	Orbital Parms.	Length of study
Willams et al,1974	Jan, Jul	5x5, L6	18k	0k	18k	18k	0k	80 days
Gates,1976	Yes (Jul)	4x5, L2	CL76	18k	18k	18k?	0k	92 days
Manabe & Hahn, 1977	Yes (Jul)	2.5x2.5,L11	CL 76	0k	0k	0k	0k	113 days
					18k	18k	0k	133 days
Hansen et al, 1984	Yes	8x10, L9	CL 81	18k	18k	18k	18k	6 years
Rind & Peteet,1985	Yes	8x10, L9	CL81	0k	18k	same as control	18k	6 years
			CL81 less 2°C	0k	18k		18k	
Rind, 1987	Yes	8x10, L9	CL81	0k	0k	0k	18k	4 years
			CL81	0k	18k (10m)	0k	18k	4 years
			CL81	0k	18k (full)	0k	18k	5 years
Kutzbach & Guetter,1986 (also Prell & Kutzbach,1987)	January July	R15,L9	CL81	18k	18k	18k	18k	450 days
				0k	15k	15k?	15k	450days (Jan) & 150 (Jul)
Manabe & Broccoli 1987 (also 1985a,b)	Yes	R15,L9	50m ML	0k	18k	0k	0k	8 years
				0k	18k	18k	0k	6 years
				18k	18k	18k	0k	8 years
Lautenschager & Hertereich,1990	Yes	T21,L16	CL81	18k	18k	18k	18k	6 years
Valdes (Pers comm)	Jan,July	T42,L19	CL81	18k	18k	18k	18k	450days
	Yes		CL81	18k	18k	18k	18k	In progress
Manabe,1991	Annual mean	R15,L9	Ocean GCM	18k	18k	18k	0k	8 years*

Resolution is either degrees (latitude by longitude) or number of spectral waves, R or T for rhomboidal or triangular truncation respectively. L indicates the number of levels. ML= mixed layer. k=1000 years before present. SST= sea surface temperature. CL76(81) =CLIMAP 1976(1981)

+CLIMAP SSTs were reconstructed for February and August; where a full annual cycle has been used, it has ben obtained by interpolation.

* 1250 years in the ocean, 34,000 years in the deep ocean.

small. The increase in land-ice produces a large cooling in the northern hemisphere only. Reducing CO_2 has the largest effect, producing a cooling in both hemispheres, but particularly in the southern hemisphere where there is a strong sea-ice/temperature feedback. Broccoli and Manabe (1987) speculate that the reduction in CO_2 was primarily responsible for the cooling in the southern hemisphere during the last glacial maximum, though they note that their model does not allow for changes in ocean circulation, which could provide an alternative explanation.

The cooling of the low latitude ocean simulated by Broccoli and Manabe (1987) is larger than that reconstructed from palaeoclimatic data (CLIMAP, 1981), particularly in the subtropics. For example, near 25°N, the zonal mean cooling is almost 3°C whereas the CLIMAP data gives a value of about 0.5°C. Manabe and Broccoli (1985b) found that changing from prescribed to model-generated cloud cover did not remove the discrepancy between simulated and CLIMAP sea surface temperatures.

The cooling deduced from data on mountains at these latitudes (Hawaii, Columbia, East Africa and New Guinea) is also much larger than one would expect from nearby sea temperature changes (Rind and Peteet, 1985). There are several estimates of the snowline descending by about 900m: assuming that the local lapse rate remains close to the mean moist adiabatic at about 4.5°C/km, this would suggest a cooling about 4°C at about 4500 m. This is consistent with a sea surface cooling of about 2°C, assuming that the temperature profile is parallel to moist adiabats (the warming at the surface becomes amplified at higher levels along a moist adiabat due to the release of latent heat) . Rind and Peteet (1985) find that a simulation with CLIMAP (1981) sea temperatures reduced everywhere by 2°C gives a better fit to the observed temperature reconstruction from tropical mountains, and a radiation balance at the top of the atmosphere which is closer to equilibrium than using the original CLIMAP (1981) values of sea temperature. The unaltered sea temperatures experiment produces only a slight decrease in the hydrological cycle in the tropics, in contrast to the reconstructed data and the results from the simulation with the lower sea temperatures. Manabe and Hahn (1977), using an older estimate of sea surface temperatures which indicated greater cooling at low latitudes (CLIMAP,1976) , also found a marked decrease in the strength of the hydrological cycle over the tropical continents.

Lautenschlager et al (1992) have attempted to simulate the circulation of the ocean at the last glacial maximum, using CLIMAP (1981) sea surface temperatures, and wind stress and water fluxes from the atmospheric simulation of Lautenschlager and Herterich (1991, see Table 2). The simulated changes are in broad agreement with palaeodata in the North Atlantic, but apparently overestimate the intensity of

overturning in the northwest Pacific. The authors speculate that this may be due in part to shortcomings in the CLIMAP (1981) sea-ice extents.

Manabe and Stouffer (1988) produced two stable equilibria in a coupled ocean-atmosphere GCM using present day boundary conditions (insolation, land ice, CO_2 and snow free land albedos). One equilibrium was similar to present, with deep water formation in a relatively warm, salty North Atlantic. In the other equilibrium, the northern North Atlantic became fresher and colder, and meridional overturning almost ceased. The authors note that the second circulation bore some similarity to that believed to have occurred during the Younger Dryas. Manabe (1991) has also reported on the results from a coupled model simulation with ice-age boundary conditions (as in Broccoli and Manabe,1987).

5. INITIATION OF ICE SHEETS

The causes of glacial/interglacial cycles have yet to be established. One possibility is that changes in the earth's orbital parameters may have led to an extension of perennial snowfields over northern North America and northwestern Europe, probably by reduced summer insolation. It has been suggested that gradual accumulation of snow may have been sufficient to build the major ice sheets, especially if augmented by other processes including temperature albedo feedbacks, the orographic effects of the growing ice sheet, ice dynamics and changes in the concentration of atmospheric CO_2.

There were particularly large changes in summer insolation in high northern latitudes around the beginning of the last glaciation, with a maximum around 125kBP, a minimum near 115kBP and a further maximum near 105kBP. Royer et al (1983), comparing a simulation for 115kBP with one for 125kBP, found that temperatures fell by several degrees and annual mean soil moisture increased in the regions of the north American and European ice sheets. However Rind et al (1989, see also Table 3) and Mitchell and Hewitt (see table 3) were unable to produce permanent snowfields in these areas through changes in orbital parameters alone. Rind et al (1989) produce stable permanent snow/ice fields in these regions only by including a 10m ice sheet, reducing CO_2 concentrations, and using full ice-age sea-ice extents and sea temperatures reductions. As Rind et al (1987) point out, this raises an awkward problem: the reductions in temperature, CO_2 and surface albedo are believed to have accompanied or even lagged the growth of the ice sheet - if so, it is hard to see how they could have contributed to the initiation of the ice sheet. They investigated possible

Table 3 Investigations of initiation of ice sheets.

Study	Sea temperatures or ocean model	CO_2 (ppm)	Land ice	Orbital parameters	Length of experiment.
Royer et al, 1983	Present	Present	Present	125k, 115k	1 year
Rind et al, 1987	Present	Present	Present	116k, Mod	All 4 years
	Mixed layer	Present	Present	116k, Mod	
	Mixed layer	Present, -70 ppm	10m	Mod	
	CL81, CL81 minus 2K	-70ppm	10m	Mod	
Mitchell and Hewitt (this paper)	Present	Present	Present	125k 115k	5 years
Oglesby, 1990	Mixed layer	100ppm, 200ppm	Present	Present	15 to 20 years
	Present	330ppm	1m snow	Present	2 years

k= 1000 years before present, Mod = 5 times the change from 114k and 116 k, CL81 = CLIMAP81 .

shortcomings in their model (including the lack of spatial resolution, which leads to lower than observed orography) and also considered additional potential forcing factors (volcanic dust, changes in solar constant), and possible re-interpretation of the palaeoclimatic data, but were unable to reconcile the simulated and reconstructed data.

Oglesby (1990) analysed the sensitivity of glaciation to reductions in CO_2 and to initial snow cover. By reducing CO_2 to 200ppm or by prescribing an initial snow depth of 1m over the areas of interest, he was able to obtain permanent snow cover over part of the region occupied by the major ice sheets. Oglesby (1990) performed additional experiments which indicated that the simulated changes are very sensitive to the fraction of the grid square assumed to be covered by snow.

Note that an increase of 1mm/day (water equivalent) in snow accumulation, or a reduction of $4Wm^{-2}$ heat available for melting, are equivalent to the growth of a 1km ice-sheet in 1000 years - it is unlikely that current models can attain that accuracy on a regional scale in high latitudes. Thus, as indicated in the introduction, this approach to investigating the initiation of the last great ice sheets should be regarded as exploratory.

6 CONCLUDING REMARKS

Numerical simulations suggest that much of the reconstructed climate change in the mid-Holocene can be accounted for by changes in the earth's orbital parameters. Simulated changes are sensitive to other factors, including the presence of ice sheets early in the period, differences in land surface characteristics due to altered vegetation, and changes in tropical sea temperatures. Better estimates of these parameters are needed to reduce the uncertainties in simulations for this period. An accurate reconstruction of temperature changes might enable some narrowing of the range of climate sensitivity produced by current models.

One of the main problems arising from the reconstruction of the last glacial maximum is the apparent discrepancy between estimates of sea temperature change in low latitudes, and the estimate from terrestrial data on nearby mountains. Current climate models suggest that the tropical sea temperatures should have been substantially lower than CLIMAP: if this is not the case, it suggests that the models are too sensitive in the tropics. This would imply that estimates of global warming due to increases in greenhouse gases have been exaggerated. On the other hand, it is hard to imagine a mechanism which would produce the large changes in lapse rate implied by the palaeodata.

If we are to attempt to understand the reasons for the last glacial period, we need to know the evolution of the extent and height of the major ice sheets more accurately. In particular, it is important to know in which order the changes in temperature, ice volume and CO_2 occurred. The effect of increased atmospheric aerosol loading has yet to be quantified.

There are many areas for improvement in current models. There is a large range of sensitivity (1.9 to 5.4°C to doubling CO_2) which is mainly due to problems in the represention of cloud and associated processes, and there is little confidence in the simulation of regional detail. Most studies to date have not taken into account changes in the ocean circulation and, as noted above, coupled ocean-atmosphere models require corrective adjustments to prevent the simulated climate from drifting unrealistically. There is considerable effort to reduce these uncertainties through observational and modelling programmes (see, for example, Houghton et al, 1990,1992). As coupled models improve, it will be possible to investigate the role of the oceans in past climate changes.

Finally, there are an increasing number of groups which are applying general circulation models to the problems of palaeoclimate. By comparing results from different models, it may be possible to identify which aspects of the simulated response are robust, and possibly more dependable, and which are dependent on the model formulation. A first step in this direction is the Palaeoclimate Modelling Intercomparison Project (PMIP) which will initially concentrate on simulations for 6kBP and the last glacial maximum.

7. ACKNOWLEDGEMENTS

I am grateful to Chris Hewitt for carrying out experiments for 6, 9, 115 and 125 kBP.

8. REFERENCES

Barnola, J.M., D. Raynaud, Y.S. Korotkevitch and C. Lorius, 1987. Vostok ice-core provides 160 000-year record of atmospheric CO_2. *Nature*, **329**, 408-414.

Berger, A.L., 1978. Long term variations of daily insolation and quaternary climatic changes, *J.Atmos.Sci.*, **35**, 2362-2367.

Bigg, G.R., D. Jiang and J.F.B.Mitchell, 1991. A quantitative assessment of the Arabian Sea sediment record using GCM simulated upwelling. *Quat.Res.*, **35**, 464-468

Boulton, G.S., G.D.Smith, A.S.Jones and J.Newsome, 1985. Glacial geology and glaciology of the last mid-latitude ice sheets. *J. Geol. Soc. London*, **142**, 447-474

Broccoli, A.J. and S.Manabe, 1987. The influence of continental ice, atmospheric CO_2 and land albedo on the climate of the last glacial maximum. *Clim. Dyn.*, **1**, 87-89

Cess, R.D., G.L.Potter, J.P.Blanchet, G.J.Boer *et al.*, 1989. Interpretation of cloud-climate feedback as produced by 14 atmospheric general circulation models. *Science*, *245*, 513-516

CLIMAP project members, 1976. The surface of the ice-age earth, *Science*, **191**, 1131-1138

CLIMAP project members, 1981. Seasonal reconstruction of the earth's surface at the last glacial maximum, map.chart. ser., MC-36, Geol. Soc. of Am., Boulder,Colo.

COHMAP members, 1988. Climatic changes of the last 18,000 years: observations and model simulations. *Science*, **241**, 1043-1052

Cullen, M.J.P., 1993. The unified model. *Meteorol. Mag.* (to appear)

Denton, G.H. and T.J.Hughes, Eds.., 1981. The last great ice sheets. Wiley, 484pp

Dixon, K.W., R.J. Stouffer, R.H.Gammon, and J.L. Bullister, 1993. Examination of a coupled ocean-atmosphere climate model using chlorofluorocarbons as transient ocean tracers (manuscript in preparation)

Folland , C.K., J. Owen, M.N.Ward and A.Coleman, 1991. Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J. of Forecasting*, **10**, 21-56.

Gallimore, R.G. and J.E.Kutzbach, 1989. Effects of soil moisture on the sensitivity of a climate model to orbital forcing at 9000 yr BP. *Climatic Change*, **14**, 175-205

Gates, W.L., 1976. Modelling the ice-age climate. *Science*, **191**, 1138-1144.

Gates W.L., J.F.B.Mitchell, G.J. Boer, U.Cubasch and V.P.Meleshko, 1992. Climate modelling, climate prediction and model validation. In *Climate Change 1992. The supplement to the IPCC Scientific Assessment* (eds J.T Houghton, B.A.Callander and S.K. Varney), 97-134. Cambridge University Press.

Gates,W.L., P.R. Rowntree and Q.-C.Zeng, 1990. Validation of climate models.In *Climatic Change: The IPCC Scientific Assessment* (eds J.T.Houghton, G.J.Jenkins and J.J.Ephraums), 93-130. Cambridge University Press.

Hansen, J.E. , A.Lacis, D.Rind, L.Russell *et al*, 1984. Climate sensitivity analysis of feedback mechanisms. In *Climate Processes and Climate Sensitivity* (ed. J.Hansen and T.Takahashi) Geophysical Monograph 29, 130-163. American Geophysical Union

Hecht, A. (ed) 1985 *Paleoclimatic analysis and modeling*, pp445, John Wiley, New York

Houghton,J.T., B.A.Callander and S.K. Varney, 1992. *Climate Change 1992. The supplement to the IPCC Scientific Assessment.* pp200. Cambridge University Press

Houghton J.T, G.J.Jenkins and J.J Ephraums(eds), 1990. *Climatic Change: The IPCC Scientific Assessment.* pp365, Cambridge University Press

Kutzbach, J.E., 1981. Monsoon climate of the early Holocene: climatic experiment using the earth's orbital parameters for 9000 years ago. *Science*, **214**, 59-61.

Kutzbach, J.E. and R.G. Gallimore, 1988. Sensitivity of a coupled atmosphere /mixed layer ocean model to changes in orbital forcing at 9000 years B.P. *J.Geophys.Res., 93*, 803-821

Kutzbach, J.E. and P.J. Guetter, 1986. The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18,000 years. *J.Atmos.Sci.*, **43**, 1726-1759

Kutzbach, J.E. and B. Otto-Bleisner, 1982. The sensitivity of the African-Asian monsoonal climate to orbital parameter changes for 9000 years B.P. in a low resolution general circulation model. *J. Atmos. Sci.*, **39**, 1177-1188

Kutzbach, J.E. and F.A.Street-Perrott, 1985. Milankovitch forcing of fluctuations in the levels of tropical lakes from 18 to 0kyr B.P.. *Nature*, **317**, 130-134

Lautenschlager,M. and K.Herterich, 1990. Atmospheric response to ice age conditions: climatology near the earth's surface. *J.Geophys.Res.*, ***95***, 22547-22557

Lautenschlager,M., U. Mikolajewicz, E. Maier-Reimer and C.Heinze, 1991. Application of ocean models for the interpretation of AGCM experiments on the climate of the last glacial maximum. Max Planck Institut fur Meteorologie, Report No. 80. Hamburg.

Liao, X., 1992. Comparison of numerical expriments using different cloud schemes and with palaeoclimatic data for 6000yr BP. D Phil Thesis, University of Oxford

Luther, M.E., J.J.O'Brien and W.L. Prell, 1990. Variability in the upwelling fields of the northwest Indian Ocean. 1. Model experiments for the past 18000 years, *Palaeoceanography*, **5**, 433-445.

Manabe, S., 1991. Studies of glacial climates by coupled ocean-atmosphere models: how useful are coupled models? In "Global changes of the past", ed R.S.Bradley, 421-448. UCAR,Boulder.

Manabe, S. and A.J.Broccoli, 1985a. The influence of continental ice sheets on the climate of an ice age. *J.Geophys.Res.*,.**90**,.2167-2190.

Manabe, S. and A.J.Broccoli, 1985b. A comparison of climate model sensitivity with data from the last glacial maximum. *J. Atmos. Sci.*, **42**, 2643-2651

Manabe, S. and D.G.Hahn, 1977. Simulation of the tropical climate of an ice age. *J. Geophys. Res.*, **82**, 3889-3991

Manabe, S. and R.J.Stouffer, 1988. Two stable equilibria of a coupled ocean-atmosphere model. *J.Clim.*, **1**, 841-866

Mitchell, J.F.B., 1977. The effect on climate of changing the earth's orbital parameters: two summer integrations with fixed sea surface temperatures. Met O. 20 Tech Note II/100, Meteorol.Office, Bracknell, England

Mitchell, J.F.B., 1989.. The "greenhouse effect" and climate change. *Rev. Geophys.*, **27**, 115-139

Mitchell, J.F.B., 1990. Greenhouse warming: is the mid-Holocene a good analogue? *J.Clim.*, **3**, 1177-1192

Mitchell, J.F.B., N.S.Grahame and K.J.Needham, 1988. Climate simulations for 9000 years before present: seasonal variations and effect of the Laurentide ice-sheet. *J.Geophys.Res.*, **93**, 8283-8303

Mitchell, J.F.B., C.A. Senior and W.J. Ingram, 1989. CO_2 and climate: a missing feedback? *Nature*, **341**, 132-4

Mitchell, J.F.B. and C.A. Senior, 1993. The unified model (climate): version 1:Atmosphere. Hadley Centre Internal Note 33.

Mix, A.C., W.F.Ruddiman and A.McIntyre, 1986. Late Quaternary paleoceanography of the tropical Atlantic. 1. Spatial variability of annual mean sea-surface temperatures. *Paleoceanography*, **1**, 43-66

Oglesby, R.J., 1990. Sensitivity of glaciation to initial snowcover ,CO_2, snow albedo and oceanic roughness in the NCAR CCM. *Clim.Dyn.*,**4**, 219-235

Petit-Maire, N, J. Riser, J. Fabre and D. Commelin, 1988. Le Sahara a l'Holocene: Mali 1;1000,000. CCGM. Paris : Institut G'eographique National.

Prell, W.L. and J.E. Kutzbach, 1987. Monsoon variability over the past 150,000 years. *J. Geophys. Res.*, **92**, 8411-8425

Reed, D.N., 1986. Simulation of temperatures and precipitation over eastern England by an atmopheric general circulation model, *Journal of Climatology*, **6**,233-253

Rind, D., 1987. Components of an ice age circulation. *J. Geophys. Res.*, **92**, 4241-4281

Rind, D. and D. Peteet, 1985. Terrestrial conditions at the last glacial maximum and CLIMAP sea-surface temperature estimates: are they consistent? *Quat. Res.*, **24**, 1-22

Rind, D., D.Peteet,W.Broecker, A.MacIntyre and W. Ruddiman, 1986. The impact of cold North Atlantic sea temperatures on climate: implications for the Younger Dryas cooling (11-10K). *Clim.Dyn.*, **1**, 3-33

Rind, D., D.Peteet and G.Kukla, 1989. Can Milankovitch orbital variations initiate the growth of ice sheets in a general circulation model? *J. Geophys. Res.*, **94**, 12851-12871

Rowell, D.P., C.K.Folland, K.Maskell, J.A. Owen, and M.N.Ward, 1992. Causes and predictability of Sahel rainfall variability. *Geophys. Res . Lett*, **19**, 905-908

Royer, J.F., M.Deque and P. Pestiaux, 1983. Orbital forcing of the inception of the Laurentide ice sheet, *Nature*, **304**, 43-46

Ruddiman, W.F. and A. McIntyre, 1981. Oceanic mechanisms for amplification of the 23,000-year ice-volume cycle. *Science*, **212**, 617-627

Saltzmann, B. 1990. Three basic problems of palaeclimatic modeling: a personal perspective and review. *Clim. Dyn.*, **5**, 67-78.

Sarmiento, J.L,. 1983. A simulation of bomb-tritium entry into the Atlantic ocean. *J. Phys. Oceanogr.*, **13**, 1924-1939.

Sausen R., K.Barthel and K. Hasselmann, 1988. Coupled ocean-atmosphere models with flux corrections. *Clim. Dyn.*, **2**, 154-163

Senior C.A. and J.F.B.Mitchell, 1993. CO_2 and climate: The impact of cloud parametrizations. *J.Clim.*, **6**, 393-418

Schlesinger, M.E.(ed), 1988. Proceedings of the NATO Advanced Study Institute on physically-based modelling and simulation of climate and climate change, Erice, Italy,11-23 May 1986. Series C: Mathematical and Physical Sciences, 243, Kluwer Academic

Street-Perrott, F.A., J.F.B.Mitchell, D.S.Marchand and J.S. Brunner, 1990. Milankovitch and albedo forcing of the tropical monsoons: a comparison of geological evidence and numerical simulations for 9000 yBP. *Trans . R. Soc. Edinburgh*, **81**, 407-427

Street-Perrott, F.A., and N. Roberts, 1983. Fluctuations in closed lakes as an indicator of past atmospheric circulation patterns. In *Variations in the global water budget* eds F.A. Street-Perrot, M.A. Beran and R.S. Ratcliffe, 331-345. D.Reidel, Mass.

Trenberth, K.E. (ed), 1992. Climate system modelling. Cambridge University Press, pp788

Toggweiller, J.R., K.Dixon and K.Bryan, 1988. Simulations of radiocarbon in a coarse-resolution, world ocean model. II: Distributions of bomb-produced carbon 14. *J.Geophys. Res*, **94**, 8243-8264

Williams, J., R.G.Barry and W.M.Washington, 1974. Simulation of the atmospheric circulation using the NCAR global circulation model with ice age boundary conditions. *J .App. Met.*, **13**, 305-317

CLIMATE RESEARCH TECHNICAL NOTES

CRTN 1	Oct 1990	Estimates of the sensitivity of climate to vegetation changes using the Penman-Monteith equation. P R Rowntree
CRTN 2	Oct 1990	An ocean general circulation model of the Indian Ocean for hindcasting studies. D J Carrington
CRTN 3	Oct 1990	Simulation of the tropical diurnal cycle in a climate model. D P Rowell
CRTN 4	Oct 1990	Low frequency variability of the oceans. C K Folland, A Colman, D E Parker and A Bevan
CRTN 5	Dec 1990	A comparison of 11-level General Circulation Model Simulations with observations in the East Sahel. K Maskell
CRTN 6	Dec 1990	Climate Change Prediction. J F B Mitchell and Qing-cun Zeng
CRTN 7	Jan 1991	Deforestation of Amazonia - modelling the effects of albedo change. M F Mylne and P R Rowntree
CRTN 8	Jan 1991	The role of observations in climate prediction and research. D J Carson
CRTN 9	Mar 1991	The greenhouse effect and its likely consequences for climate change. D J Carson
CRTN 10	Apr 1991	Use of wind stresses from operational N.W.P. models to force an O.G.C.M. of the Indian Ocean. D J Carrington
CRTN 11	Jun 1991	A new daily Central England Temperature series, 1772-1991. D E Parker, T P Legg and C K Folland
CRTN 12	Jul 1991	Causes and predictability of Sahel rainfall variability. D P Rowell, C K Folland, K Maskell, J A Owen, M N Ward
CRTN 13	Jul 1991	Modelling changes in climate due to enhanced CO_2, the role of atmospheric dynamics, cloud and moisture. C A Senior, J F B Mitchell, H Le Treut and Z-X Li

CLIMATE RESEARCH TECHNICAL NOTES

CRTN 14	Sep 1991	Sea temperature bucket models used to correct historical SST data in the Meteorological Office. C K Folland
CRTN 15	Aug 1991	Modelling climate change, and some potential effects on agriculture in the U.K. P R Rowntree, B A Callander and J Cochrane
CRTN 16	Aug 1991	The Boreal Forests and Climate G Thomas and P R Rowntree
CRTN 17	Aug 1991	Development of a Stratosphere-Troposphere Data Assimilation System. R Swinbank.
CRTN 18	Sep 1991	A study of asynchronous coupling using a simple climate model. M K Davey.
CRTN 19	Sep 1991	The Oceanic Carbon Cycle. N K Taylor
CRTN 20	Nov 1991	Worldwide ocean-atmosphere surface fields in Sahel wet and dry years using provisionally corrected surface wind data. M N Ward
CRTN 21	Dec 1991	Coupled tropical ocean global atmosphere models at the UKMO. M K Davey, C Gordon, S Ineson and S Lawrence
CRTN 22	Dec 1991	Empirical parameterisation of tropical ocean-atmosphere coupling. M Allen, M K Davey, D L T Anderson and P D Killworth
CRTN 23	Feb 1992	The temporal evolution of Equatorial Currents in the Indian Ocean D L T Anderson, D J Carrington, R A Corry and C Gordon
CRTN 24	Feb 1992	Stratospheric analyses provided by the U.K. Meteorological Office M J Bailey, A O'Neill and V D Pope
CRTN 25	Feb 1992	Modelling interannual variability in the Indian Ocean using momentum fluxes from the UKMO and ECMWF operational weather analyses. D L T Anderson and D J Carrington

CLIMATE RESEARCH TECHNICAL NOTES

CRTN 26	Mar 1992	A GCM simulation of the impact of Amazonian deforestation on climate using an improved canopy representation J Lean and P R Rowntree
CRTN 27	Mar 1992	The parameterization of rainfall interception in GCMs A J Dolman and D Gregory
CRTN 28	Jun 1992	Development of worldwide marine data eigenvectors since 1985 A W Colman
CRTN 29	Jun 1992	A tropical ocean model with reduced physics M K Davey
CRTN 30	Jun 1992	International Temperature Workshop, Boulder, Colorado, USA, 16 January 1992 Edited by D E Parker
CRTN 31	Jul 1992	Simulation of clear-sky outgoing longwave radiation over the oceans using operational analyses A Slingo and M J Webb
CRTN 32	Sep 1992	A prediction of the transient response of climate J M Murphy
CRTN 33	Nov 1992	LEPS scores for assessing climate model simulations and long-range forecasts. C K Folland
CRTN 34	Jan 1993	Stratospheric Data Assimilation System Guide Editer: R Swinbank
CRTN 35	Mar 1993	A Stratosphere-Troposphere Data Assimilation System R Swinbank and A O'Neill
CRTN 36	Mar 1993	Validation of hydrological schemes for climate models against catchment data. P R Rowntree and J Lean
CRTN 37	Apr 1993	Modelling of palaeoclimates: Examples from the recent past J F B Mitchell